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1 **The influence of long- and short-term volcanic strain on aquifer pressure: a**
2 **case study from Soufrière Hills Volcano, Montserrat (W.I.)**

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20 **Summary**

21 Aquifers are poroelastic bodies that respond to strain by changes in pore pressure. Crustal
22 deformation due to volcanic processes induces pore pressure variations that are mirrored in well
23 water levels. Here, we investigate water level changes in the Belham valley on Montserrat over
24 the course of two years (2004-2006). Using finite element analysis, we simulate crustal
25 deformation due to different volcanic strain sources and the dynamic poroelastic aquifer response.
26 While some additional hydrological drivers cannot be excluded, we suggest that a poroelastic
27 strain response of the aquifer system in the Belham valley is a possible explanation for the observed
28 water level changes. According to our simulations, the shallow Belham aquifer responds to a
29 steadily increasing sediment load due to repeated lahar sedimentation in the valley with rising
30 aquifer pressures. A wholesale dome collapse in May 2006 on the other hand induced dilatational
31 strain and thereby a short-term water level drop in a deeper-seated aquifer, which caused
32 groundwater leakage from the Belham aquifer and thereby induced a delayed water level fall in
33 the wells. The system thus responded to both gradual and rapid transient strain associated with the
34 eruption of Soufrière Hills Volcano (Montserrat).

35 This case study gives field evidence for theoretical predictions on volcanic drivers behind
36 hydrological transients, demonstrating the potential of hydrological data for volcano monitoring.
37 Interrogation of such data can provide valuable constraints on stress evolution in volcanic systems
38 and therefore complement other monitoring systems. The presented models and inferred results
39 are conceptually applicable to volcanic areas worldwide.

40 **Keywords:** Hydrology, Numerical modelling, Volcano monitoring, Fracture and flow,
41 Permeability and porosity, Transient deformation

42

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60 Government of Montserrat for sharing mean sea level pressure data.

61

62

63 **1 Introduction**

64 Hydrological systems can respond to or modify the expression of magmatic processes or
65 become an agent of volcanic unrest themselves, resulting in observable transients in volcano and
66 hydrological monitoring systems (e.g., Rouwet et al., 2014, Newhall et al., 2001, Jasim et al.,
67 2018). If we understand the underlying mechanisms, we can integrate hydrological observations
68 with other volcano monitoring systems to get a more complete picture of volcanic processes and
69 related hazards and their interactions with the local hydrology.

70 Changes in water levels in wells, springs, and lakes have been reported in association with
71 eruptions at many volcanoes of different kinds and in various settings (e.g. Newhall et al., 2001,
72 and references therein). Several volcanic drivers have been proposed as causes for these changes,
73 including the injection of fluids derived from magmatic degassing into the aquifer (Capasso et al.,
74 2014) or changes in groundwater flow patterns e.g. through the opening and closure of fractures
75 (Hautmann et al., 2010, Hurwitz and Johnston, 2003).

76 An often-proposed mechanism behind changes in well water levels are strain-induced
77 changes in pore pressure. Aquifers are poroelastic media, i.e. the solid matrix behaves elastically,
78 but is coupled to the pore fluid and its flow. Poroelastic media react to applied volumetric strain
79 like a confined sponge: pore pressure rises under compression and falls during dilatation owing to
80 the decrease and increase in pore space, respectively (Wang, 2000). Conversely, the presence of
81 pore fluids also affects the deformation of the solid matrix and variations in pore pressure deform
82 the surrounding matrix (Wang, 2000). Changes in water levels due to pore pressure changes
83 resulting from seismic strains before, during or after earthquakes have been widely observed (e.g.
84 Roeloffs, 1996, Shibata et al., 2010, Kopylova and Boldina, 2012, Takahashi et al., 2012, Brodsky
85 et al., 2003).

86 Previous work on poroelastic processes at volcanoes almost exclusively considered ground
87 deformation induced by pressure changes in hydrothermal systems, a process commonly studied
88 with one-way coupled numerical models (e.g. Todesco et al., 2004, Fournier and Chardot, 2012,
89 Coco et al., 2016). Less studied are poroelastic responses of aquifers to volcanic processes such as
90 reservoir pressure changes, which frequently cause stress changes in the surrounding crust. Some
91 attempts at interpreting pre-, syn- and post-eruptive water level changes in the light of volcanic
92 strain have been performed for example at Usu and Meakan-dake volcano in Japan (e.g. Yokoyama
93 and Seino, 2000, Takahashi et al., 2012) and Krafla volcano in Iceland (Stefansson, 1981). By
94 tracing water level changes in response to known strain excitations, the strain sensitivity, i.e. the
95 well level change per unit applied strain, of an aquifer can be quantified. This method is common
96 in studies of seismically-induced water level changes and was applied by several volcanological
97 studies (e.g. Matsumoto et al., 2002, Kopylova and Boldina, 2012, Shibata et al., 2010) in order to
98 derive crustal strain from observed strain-induced water level changes.

99 Only a few, very simple poroelastic problems have been solved analytically, for example
100 the compression of a homogeneous, water-saturated block (e.g. Rice and Cleary, 1976). For more
101 complex problems, e.g. with heterogeneous media and/or more sophisticated deformation sources,
102 and for the investigation of time-dependent fluid-flow effects, a numerical approach is necessary.
103 Strehlow et al. (2016) present the first fully-coupled, time-dependent numerical models for
104 poroelastic water level changes at volcanoes. Using finite element analysis, they simulate crustal
105 deformation accompanying magma chamber pressurization and the resulting significant hydraulic
106 head changes as well as flow through the porous aquifer.

107 Following the theoretical analysis of Strehlow et al (2016) that includes a detailed
108 sensitivity analysis, this paper provides the first step towards application of these models in

109 volcano monitoring. After an introduction to the study site, i.e. the island of Montserrat, West
110 Indies, and its volcanological and hydrological system, we report volcanic activity and observed
111 water level changes in 2004-2006. This period was chosen as the focus of this study since both
112 hydrological and volcano monitoring data are available, and the time frame covers an intra-
113 eruptive quiescent period from July 2003 to July 2005, the renewal of lava extrusion in August
114 2005 and the formation and partial explosive destruction of a new dome (Odbert et al., 2014a). We
115 suggest that these water level variations are caused by volcanic strain and present results of
116 numerical models that are applied to test our hypotheses. Limits and implications of these results
117 are discussed, and we finally conclude that well level changes on Montserrat are indeed at least
118 strongly affected by volcanic processes acting on different time-scales.

119 **2 Setting**

120 2.1 Geology of Montserrat

121 Montserrat is a small (sized 16 km x10 km) volcanic island located in the north of the
122 Lesser Antilles inner volcanic arc (Fig. 1). The island consists of three volcanic centres that
123 decrease in age from north to south, i.e., from the extinct Silver Hills and Centre Hills to the
124 youngest Soufrière Hills Volcano.

125 The northernmost volcanic complex Silver Hills (SH) was active 2.6 to 1.2 million years
126 ago, while the Centre Hills (CH) were erupting about 950 to 550k years ago (Harford et al., 2002).
127 The remnants of the old andesitic volcanic domes from the extinct volcanic centres SH and CH
128 are today dominating surface expressions on the island. Their flanks are surrounded by aprons of
129 volcanoclastics, which are predominantly pyroclastic flow deposits, with lesser amounts of pumice-
130 and-ash-flow, pumice-fall, lahar, debris-avalanche and fluvial deposits (Shalev et al., 2010;

131 Hautmann et al., 2013). The volcanoclastic beds are intersected by radially incised valleys that are
132 formed by run-off from heavy rainfall (Hemmings et al., 2015a).

133 The active Soufrière Hills Volcano (SHV) in the south of Montserrat is an andesitic dome-
134 building volcano that first started erupting 175k years ago (Harford et al., 2002). After an
135 approximately 370 years period of dormancy (Kokelaar, 2002), renewed activity began in 1995.
136 Since then, SHV has undergone various cycles of lava dome extrusions and discrete Vulcanian
137 explosions alternating with phases of volcanic quiescence and cessation of dome growth (i.e.,
138 Odbert et al., 2014b). Episodes of surface activity are typically related with ground deflation, while
139 periods of quiescence are associated with a re-pressurization of the subsurface magmatic system
140 and a consequent ground inflation. After a partial dome collapse in 2010, which marked the last
141 major activity to date, the volcano entered a new episode of continuous surface inflation that is
142 still ongoing at time of this writing.

143 The two volcanic complexes CH and SHV are geologically separated by the ESE trending
144 Belham Valley fault (BVF). The BVF is part of a larger fracture system that connects the
145 Bouillante-Montserrat graben between Guadeloupe and Montserrat in the east with the Montserrat-
146 Havers Fault System that extends to the west of Montserrat (Feuillet et al., 2010). The fault appears
147 to be currently inactive: there is an absence of earthquakes in the historic record and repeated
148 electronic distance measurements indicate a lack of movements along the fault (MVO, pers.
149 comm.). The Belham Valley, which is the morphological expression of the BVF, is a major
150 drainage channel that has been filled with >12 m lahar and pyroclastic deposits since the beginning
151 of the SHV eruption (Froude, 2015).

152 2.2 The SHV magmatic system

153 The analysis of geodetic data from different volcanic activity phases of SHV allowed for
154 the identification and characterization of three connected pressure sources in the crust (e.g. Odbert
155 et al., 2014b). Assembling these inferred sources gives a model of a magma plumbing system that
156 consists of two vertically-stacked magma chambers that link to the surface via a dyke-conduit
157 feeder system (i.e., Hautmann et al., 2013, 2014). The lower magma chamber (LMC) has been
158 found to be prolate with its centroidal source depth located at ~13 km bsl (Hautmann et al., 2010),
159 while the upper magma chamber (UMC) has been inferred to be spherical, centred at ~6 km depth
160 (Voight et al., 2006). Independent studies on magma flow dynamics, seismic tomography and
161 strain data inversion constrained the UMC:LMC magma chamber volume ratio to 1:3, whereof the
162 LMC is estimated to have a size of 8 km³ (Paulatto et al., 2012; Hautmann et al., 2013; Melnik and
163 Costa, 2014). A NW-SE trending dyke that opens at ~1 - 1.5 km depth into a small cylindrical
164 conduit connects the UMC with the surface (Mattioli et al., 1998; Costa et al., 2007; Hautmann et
165 al., 2009; Linde et al., 2010; Gottsmann et al., 2011).

166 2.3 Hydrology and the Belham aquifer

167 The hydrological system of Montserrat is characterized by high-yielding springs on the
168 upper flanks of the extinct CH and low-lying coastal aquifers in the volcanoclastic aprons
169 (Hemmings et al., 2015a). High infiltration rates limit the surface run-off on Montserrat and
170 streams exist only during and shortly after intense rainfall events, when water flows through deep
171 valleys towards the sea. Rainfall occurs throughout the year but follows a clear seasonality. The
172 wet season ranges from July to November with the highest precipitation values in September to
173 November, and the dry season spans from February to April (Hemmings et al., 2015a). Even
174 though infiltration rates are high, high rates of interception and evapotranspiration significantly
175 reduce the percentage of precipitation that reaches the ground to recharge. Additionally, there is

176 significant run-off during the very large rainfall events, which further limits recharge. Hemmings
177 et al. (2015a) predict an annual recharge of 10–20% of annual rainfall with a pronounced
178 seasonality, as 70% of the island's recharge occurs between July and December. Recharge rates
179 vary significantly both temporally and spatially. Transpiration is especially high in the forested
180 areas, which limits recharge, while infiltration and recharge rates are high on fresh volcanic
181 deposits (Hemmings et al., 2015a). Simulated estimates of mean (whole island) recharge rates for
182 individual months range from 7.2 (June) to 55.8 mm/month (October) for the whole island,
183 increasing to 217 mm/month in the southern part of the island (Hemmings et al., 2015a).

184 The Belham valley is the broadest drainage channel on the island with a catchment area of
185 about 16 km² (Froude, 2015). It hosts a confined aquifer in reworked gravels and alluvial deposits
186 at depths between 15 and 38 m below sea level (bsl) (Hemmings et al., 2015a) (hereafter called
187 the "Belham aquifer"). Hemmings et al. (2015a, 2015b) also propose a second, deeper and warmer
188 aquifer that feeds the springs of Centre Hills and the Belham aquifer by upflow through a fault and
189 fracture network. The Belham aquifer is confined by a thin cap (about 1 m) of low permeability
190 clay and covered by several meters of lahar deposits. In 2003, previously existing wells, drilled for
191 groundwater development in the Belham valley, were buried and lost under fill accumulations
192 from lahars. In order to regain back-up water supply, HydroSource Associates drilled three new
193 wells in 2004 down to depths of 79 m below ground surface (bgs) (MBV 1), 59 m bgs (MBV 2)
194 and 110 m bgs (test well), respectively (HydroSource 2004). The test well is a pure observation
195 well, while MBV 1 and MBV 2 serve as backup water sources, but are not normally pumped for
196 supply purposes and never pumped during periods of heightened volcanic activity. All wells are
197 hydraulically connected, evidenced by pumping tests conducted by HydroSource Associates.
198 These tests also revealed that the tapped Belham aquifer is highly permeable (10-10 m²)

199 (HydroSource, 2004). The water is comparatively warm (31° C) and the aquifer is somewhat
200 buffered from atmospheric and seasonal fluctuations in recharge (Hemmings et al., 2015b),
201 although the spatial and temporal uncertainties regarding recharge are significant and preclude
202 well-founded statements regarding the influence of recharge on the wells. In 2012, we operated a
203 continuously recording pressure logger (sample frequency 0.003 Hz) in the test well of Belham
204 valley (see online resource 1), which allows the investigation of possible short-term effects in the
205 aquifer. These data show that the tidal signal in the aquifer is smaller than the general noise in the
206 data. Tidal responses can only cause hydraulic head changes < 0.5 cm and were therefore neglected
207 in this study. Comparison with rainfall data showed furthermore that the system is largely buffered
208 from short-term meteoric events.

209

210 **3 Observations 2004 - 2006**

211 3.1. Aquifer head changes

212 The focus of this study is the analysis, modelling and interpretation of water level data
213 recorded between November 2004 and December 2006 in the three wells tapping the Belham
214 aquifer. Provided that weather conditions and volcanic activity allowed access to the valley, water
215 levels in the wells were measured with a portable tape dipper on a roughly weekly basis, with only
216 a single larger gap in the data, between 14th of June and 10th of August 2005. The recorded water
217 level changes (Fig. 2a) indicate three features:

218 (i) Between November 2004 and May 2006 there is an annual periodicity with a
219 pronounced increase of water levels (0.4 - 0.8 m) between October and March and a slight decrease
220 (~10 cm) of water levels during the summer months.

221 (ii) Over the entire observation episode, water levels increased by almost 2 m. Particularly
222 after May 2006, this increase follows a general linear trend and even swamps the yearly periodic
223 signal. This increase seems to have continued in some form after the observation period, because
224 one of the wells became flowing artesian in 2011 and has been discharging water into the valley
225 until (at least) 2014 (Hemmings et al. 2015a; Hemmings et al. 2012).

226 (iii) A notable decrease in water levels by 10 cm is documented between the 25th and 31st
227 of May 2006. Occasionally, one of the three wells shows a short deviation from the overall trend
228 but the decrease at the end of May 2006 is the only short-term signal visible in all three wells. We
229 therefore assume it to be a real effect even though the water level fall is based on just one data
230 point in time.

231 The Belham aquifer was saturated during the whole period of observations: using the initial
232 depth to water level in the wells (between 3.2 m at MBV2 and 6.63 m at the test well) and the
233 elevation of the wells (about 40 m asl), we can infer that the well water levels were always well
234 above the upper boundary (38 m bsl) of the aquifer.

235 3.2. Barometric pressures and rainfall

236 Barometric pressure variations affect well water levels in a complex way and the
237 barometric efficiency and its variation with time depend on aquifer properties and hydrological
238 conditions. The general observation is that in response to an atmospheric pressure drop, well levels
239 rise and vice versa (Rojstaczer and Agnew 1989; Rasmussen and Crawford 1997). We therefore
240 compare the hydrologic data with mean sea level pressure changes (Fig. 2b) that were recorded
241 simultaneously at Montserrat's airport (168 m asl), which is located between SH and CH (data
242 provided by Karen Pascal, owned by the Government of Montserrat). Monthly averaged mean sea
243 level pressures generally display an annual depression between September and November,

244 correlating with the months with highest precipitation values. Rainfall data collected during the
245 same time interval by the Montserrat Utilities at the Hope rain gauge document the seasonality
246 typical for Montserrat, with largest precipitation values between September and November (Fig.
247 2b).

248 In order to test the relation between meteorological parameters and the water levels in the
249 Belham valley, we calculated correlation coefficients of the different data sets, which are shown
250 in Table 1. Cumulative precipitation and monthly averaged mean sea level pressure (before May
251 2006) both show a good correlation with the well water level. Daily measured precipitation and
252 mean sea level pressure data do not correlate with water levels.

253

254 3.3 Volcanic activity

255 The studied monitoring window was characterized by a switch in volcanic activity from magma
256 chamber inflation and dome-growth cessation (representing the second pause since the beginning
257 of the eruption in 1995) to magma chamber deflation and lava extrusion at the surface (activity
258 episode 3 since 1995). The pause of volcanic activity began in July 2003 after a major dome
259 collapse at SHV and ended in April 2005 with the onset of a series of phreatic and later Vulcanian
260 eruptions, which were precursory to the beginning of a new dome growth phase in August 2005
261 (Wadge et al., 2014). Since then, the lava extrusion rate increased exponentially (Fig. 2a) until a
262 major collapse on 20th May 2006 removed all dome material that was extruded in Activity Phase
263 3 and remnants of the earlier Activity Phase 2. The collapse lasted 3 h and involved a total volume
264 of 97×10^6 m³ (dense rock equivalent) (Wadge et al., 2014). The May 2006 collapse was in its
265 mechanisms (direction, related surges, explosions) similar to the major collapse in July 2003,
266 which marked the end of Activity Phase 2, but the May 2006 event was smaller by a factor of 2

267 with regard to dome volume removal (Voight et al., 2006). Dome growth resumed subsequently
268 after the collapse in May 2006 and remained steady at a high rate until early 2007.

269

270 **4 Hypotheses**

271 We propose that the observed water level changes in the Belham aquifer can be caused, or
272 are at least heavily affected, by volcanic processes that induce both long-term and short-term stress
273 changes leading to poroelastic pore pressure changes in the aquifer. This study focuses on the
274 investigation of the aquifer response to volcanic strain and demonstrates that this is an essential
275 component of the drivers behind observed water level changes. This does not rule out the potential
276 contribution of purely hydrological mechanisms (e.g. hydrological recharge and barometric
277 responses) to observed water level signals, although these are ignored in the simulations.

278 Regarding the potential barometric contributions to water level signals, according to Freeze
279 and Cherry (1979), barometric efficiency of confined aquifers usually falls in the range of -0.2 to
280 -0.8 cm/mbar. During the observation period, the day-to-day variation in mean sea level pressure
281 on Montserrat was less than 5 mbar (Fig. 2b), which translates into water level changes between 1
282 and 4 cm. We can dismiss atmospheric pressure changes as a cause for the long-term water level
283 increase of about 2 m. The short-term water level fall would require a short-lived, high pressure
284 weather system on 31 May 2006, which cannot be seen in the mean sea level pressure data set, and
285 thus this abrupt water-level change also cannot be attributed to an atmospheric pressure response.
286 Although atmospheric pressure follows a seasonal trend and thus might play a role in the observed
287 periodicity of the water levels, the seasonal signal in the water level data is significantly larger
288 than the expected response to atmospheric pressure variations.

289 Estimation of the potential contribution from recharge variations is less straight-forward.
290 Recharge in the Belham valley is seasonal and may thus contribute to the seasonal signal in the
291 water level data. Climate-driven seasonal and interannual recharge variations on Montserrat are
292 also potentially coupled with recharge changes induced by the destruction of vegetation (and
293 therefore, transpiration potential) by volcanic activity (Hemmings et al 2015a). While precipitation
294 did not significantly change (Fig. 2b), recharge modelling by Hemmings et al. (2015a) showed
295 that, due to lower evapotranspiration and higher infiltration rates, recharge on Montserrat is almost
296 5 times higher on bare soils and fresh volcanic deposits than on forested regions. Due to the intense
297 volcanic activity of the last 20 years, vegetation has been damaged or destroyed over large areas
298 of the island and this likely increased the hydrological recharge. Therefore, a general increase in
299 recharge in the catchment area could be a contributing mechanism behind the long-term trend of
300 rising water levels in the Belham valley. To estimate this effect would require piezometric data
301 with the same hydrological regime as the Belham aquifer but not influenced by the volcano;
302 unfortunately these data do not exist. Due to the significant temporal and spatial variations and
303 unknowns regarding the hydrological dynamics of the system (e.g. the hydraulic properties of the
304 aquifer, its connectivity to recharge locations, as well as the exact catchment area, are unknown),
305 the explicit quantification of this contribution is beyond the scope of this paper and will be
306 addressed in a hydrologically focussed study at later stage.

307 Here, we explore scenarios related to volcanic activity causing the observed signals. The
308 most prominent long-term stress changes on Montserrat are due to the inflation-deflation cycles of
309 SHV, however, water levels increased during both repose and lava dome extrusion phases (Fig.
310 2a). Therefore, the overall trend of water level change is not related to reservoir pressure changes
311 and lava extrusion at SHV. A short-term signal in water levels could be linked to seismic stress.

312 However, the seismic record for the time period of the observed water level fall in May 2006 does
313 not indicate either unusually low or high activity (e.g. Loughlin et al., 2006).

314 Instead, we propose lahar sedimentation and a dome collapse as driving forces behind
315 poroelastic water level changes in the Belham valley.

316 4.1 Hypothesis 1: Lahar loading

317 The Belham valley was inundated by numerous lahars after the onset of the eruption, during both
318 repose and extrusion phases (Froude, 2015; Alexander et al., 2010; Donnelly, 2015; Barclay et al.,
319 2007). Ten and 23 lahars are confirmed or regarded as very likely to have entered the valley in
320 2005 and 2006, respectively (Froude, 2015). The lahars are caused almost exclusively by rainfall,
321 which mobilises loose volcanic deposits in the upper catchment and transports material
322 downstream towards the coast (Barclay et al., 2007). Lahar occurrence is therefore much higher in
323 the rainy season compared to the dry season, with October showing the highest lahar incidence of
324 3.4 lahar days per month and February having the least lahars with 0.3 lahar days per month
325 (averaged from the lahar record 1995-2013) (Froude, 2015). Since flow volume data are not
326 available, Froude (2015) distinguished lahar sizes based on their duration and flow width: 6% of
327 all lahars occupied most of the valley floor and lasted for over 24 hours, 29% occupied more than
328 50% of the valley floor and persisted for more than 12 hours, while 65% were confined to a single
329 channel and lasted just a few hours. The lahar sediment deposition has led to an average sediment
330 aggradation of 0.4 m per year, dramatic geomorphic changes - including a significant seaward
331 movement of the shoreline - and deep burial of houses and infrastructure (Froude, 2015). An
332 increase in the sediment burden affects underlying water-saturated porous media by increasing the
333 pore pressure (e.g. Boutt, 2010). We therefore propose that the long-term water level increase in
334 the Belham aquifer is caused by the increasing sediment load due to repeated lahars (*hypothesis*

335 1). Since the lahars are induced by rainfall, cumulative rainfall data are related to cumulative lahar
336 loading, and the good correlation of cumulative rainfall data with well observations (Tab. 1)
337 supports this hypothesis. The periodicity of water levels can therefore be linked to the higher
338 number of lahars during the rainy season and the absence of rainfall-induced lahars during the dry
339 season.

340 4.2 Hypothesis 2: Dome collapse

341 The timing of the single notable water level decrease (31st of May 2006) is significant, since a
342 wholesale lava dome collapse occurred on 20 May 2006 (Fig. 2a), removing 97 Mm³ of material
343 from the summit (Ryan et al. 2010; Loughlin et al. 2010; Trofimovs et al. 2012). The dome material
344 travelled down the volcano's eastern flank (i.e. away from the wells) and most was deposited
345 offshore. We propose that the short-term water level decrease in the Belham aquifer is related to
346 strain induced by the dome collapse. A direct poroelastic response, however, would be immediate.
347 Additionally, due to the distance of the strain source, a direct elastic response in the shallow,
348 unconsolidated Belham aquifer will be small or even negligible, because sufficient strain-coupling
349 requires more competent lithologies at these distances. Therefore, we suggest that the collapse
350 induced a significant hydraulic head drop in a second, deeper-seated aquifer (in line with
351 suggestions by Hemmings et al. (2015a, 2015b)), which is more competent and connected to the
352 Belham aquifer by vertical fractures. This would cause porous and/or fracture flow between the
353 aquifers and thus explain the time delay between dome collapse and water level fall in the Belham
354 aquifer. The Belham valley is a fault zone with a fracture network that can facilitate such
355 connections (Hautmann et al., 2010; Feuillet et al., 2010; Kenedi et al., 2010). Furthermore,
356 Bouguer anomalies suggest a significant shift in rock density in the Belham valley at depths greater
357 than 600 m (Hautmann et al., 2013) and, while recent volcanic deposits from SHV are not usually

358 of a sufficient Young's Modulus (Young and Gottsmann, 2015), the compaction and hydrothermal
359 consolidation by geothermal fluid input (proposed by Jones et al. (2010)) could increase the
360 stiffness of a deeper-seated aquifer. The connection between the aquifers may be intermittent -
361 fractures can form and widen (increasing permeability) due to seismic or other perturbances and
362 become sealed again with time (e.g. Montgomery and Manga, 2003, Rojstaczer and Wolf, 1992,
363 Elkhoury et al., 2006, Shi et al., 2015, Geballe et al., 2011).

364 Crustal strain due to dome collapse has two origins. The first is unloading of the summit surface
365 due to the removal of weight. The second strain source is magma chamber inflation as a result of
366 surface unloading. Recorded strain data associated with the dome collapse in July 2003 were
367 interpreted by Voight et al. (2006) to indicate inflation of SHV's shallower magma chamber
368 shortly after the collapse (see also Chen et al., 2018). Voight et al. (2006) propose that the drop in
369 lithostatic load due to the collapse led to vesiculation of the resident magma, causing a rapid (about
370 4 h) build-up of pressure after the collapse. We suggest that similar processes were associated with
371 the 2006 dome collapse and that the Belham aquifer system responded to these volcanic strains
372 (i.e. unloading from the dome and short-term increase of chamber pressure) as a poroelastic
373 medium, which in turn can explain the observed water level changes (*hypothesis 2*).

374 Both hypotheses are illustrated in a conceptual model in Fig. 3. In the following we test
375 these hypotheses and explore the associated mechanisms by simulating the scenarios in numerical
376 models that account for poroelastic deformation of the aquifer.

377

378 5 Methods

379 We apply poroelastic models developed by Strehlow et al. (2015) in the finite element analysis
380 software COMSOL Multiphysics (version 5.1), benchmarked against a known analytical solution
381 for a poroelastic problem (see online resource 2). While some problems presented in this study
382 could be solved analytically, we implemented finite element models because they allow us to
383 include stratigraphy and topography easily. These time-dependent, 2D-axisymmetric models solve
384 fully-coupled solid mechanics and porous flow equations to simulate crustal deformation and the
385 resulting pressure changes and fluid flow in a water-saturated, confined aquifer. Additional
386 hydrological sources and sinks are neglected. The generic models are adapted to incorporate
387 available information on aquifer properties, strain sources and surrounding lithologies on
388 Montserrat. We assume linear elastic and poroelastic behaviour of the surrounding crust and the
389 aquifer domain, respectively. The simulated flow is isothermal. Chosen material properties are
390 based on literature (Freeze and Cherry, 1979, Fetter, 1994, Wang, 2000, Gercek, 2007,
391 Gudmundsson, 2011, Adam and Otheim, 2013, geotechdata.info, 2013, Sevilla et al., 2010, Young
392 and Gottsmann, 2015, Hemmings et al., 2015a). All reference values of input parameters are listed
393 in Table 2; more details and ranges for parameter sweeps are provided in online resource 3. We
394 run the simulations with both aquifer properties representing the Belham aquifer, and for a stiffer,
395 less permeable aquifer (model acronyms with an “s”-subscript), representing lava or a more
396 consolidated pyroclastic deposit.

397 The initial pore pressure in the aquifer is set as hydrostatic. The model solution gives solid
398 displacement \mathbf{u} and fluid pore pressure p_f . For comparison with the observed water table changes,
399 we present model results as the hydraulic head:
$$h = \frac{p_f}{\rho_f \times g} - z \quad (1) \quad (\rho_f: \text{water density, } g:$$

400 gravitational acceleration, z : depth coordinate), which is proportional to pore pressure and
401 represents the maximum water level change in a small diameter well in a confined aquifer. As the
402 wells move with the ground, vertical ground displacement is subtracted from this hydraulic head
403 change to obtain the relative water level change that would be measured in the wells.

404

405 5.1 Models

406 Our hypotheses involve a stacked, two-aquifer system in the Belham valley with a (possibly
407 intermittent) connection through pores and/or fracture flow. Since the type and diffusivity of this
408 connection is unknown, and in order to be able to independently resolve and understand
409 hydrological processes in response to different strain sources in detail, we set up two suites of
410 models. The first set of models simulates the effect of lahar loading on an aquifer ("lahar models",
411 acronym L). The second set simulates aquifer response to strain due to a dome collapse ("collapse
412 models", acronym COL). The more complex response of a stacked aquifer system significantly
413 limits the efficiency of the models and is out of the scope of this study and left for future work.
414 The presented models assume a confined, water-saturated aquifer that does not undergo significant
415 temperature changes during the period of interest (for more details see Strehlow et al. (2015)).
416 These are valid assumptions at least for the local vicinity of the wells, since temperatures have

417 been approximately constant in the past (Hemmings et al., 2015b) and the Belham aquifer was
418 saturated during the whole period of observations.

419 5.1 Lahar models

420 To investigate the influence of sediment loading and test hypothesis 1 ("lahar models",
421 acronym L), we define the system as shown in Fig. 4a consisting of a linear elastic solid block
422 with an embedded shallow poroelastic aquifer that is water-saturated and confined. The duration
423 of the time-dependent simulation is 3 years. We apply a time-dependent upper boundary load to
424 the surface that corresponds to an annual sediment aggradation a (reference value of 0.4 m/a):

$$425 \quad \Delta P_{lahar}(t) = \rho_{sediment} \times g \times a \times t \quad (2)$$

426 with $\rho_{sediment}$ the density of the deposited sediment and t time. In the first suite of models, we
427 investigate a (temporally) linear loading. Since lahar frequency is coupled with rainfall (Barclay
428 et al. 2007), we then adapt the load such that it mirrors the inter-seasonal variation in rainfall on
429 Montserrat (model acronym Lc): no load is applied in the dry season (February to April), while
430 the load during September to November is twice as high as in the remaining months, following the
431 seasonality of rainfall outlined in Hemmings et al. (2015a). The reference lahar model with cyclic
432 loading (Lc) can be found as an .mph file in online resource 4.

433 5.2 Collapse models

434 Fig. 4b shows the geometry used to test hypothesis 2 ("collapse models", acronym COL),
435 which includes an approximation of the volcano's topography and its magmatic plumbing system
436 (magma chambers are represented as cavities in the domain). The aquifer in these simulations does
437 not cover the whole domain but starts at a variable lateral distance from the volcanic summit: as a

438 reference value, we chose the distance from the vent to the Belham valley fault. We simulate
 439 crustal deformation associated with dome collapse and growth, which resumed immediately
 440 afterwards, by assigning boundary loads that represent the two sources contributing to generated
 441 strain: the loading and unloading of the summit's surface by the weight of the dome material, and
 442 the inflation and deflation of the volcano's plumbing system. The duration of the time-dependent
 443 simulation is 200 days.

444 *Loading of the summit:* A total rock volume of $V_{Dome}=97 \text{ Mm}^3$ (dense rock equivalent) was
 445 removed from the summit during the May 2006 dome collapse (Loughlin et al. 2010). Assuming
 446 a cylindrical dome with a base radius r_{Dome} of 0.5 km, we can define the unloading function as:

$$447 \quad \Delta P_{collapse} = -\rho_{lava} \times g \times h_{av} = -2600 \frac{kg}{m^3} \times g \times \frac{V_{Dome}}{\pi \times r_{Dome}^2} = -3.15 \text{ MPa} \quad (3)$$

448 with ρ_{Lava} the average density of extruded dome lava (2600 kg/m³) and h_{av} the average height of
 449 the dome at the time of collapse. Most of the dome collapse occurred in less than 1 hour (Loughlin
 450 et al., 2010), so we ramp up this collapse function linearly from 0 to 100% over the duration of 1
 451 hour.

452 Using the average extrusion rate $\frac{V}{t}$ of the dome in the period after the dome collapse (which
 453 was roughly constant – see Fig. 2a), we can also define the time-dependent loading function for
 454 the summit during dome growth using the dome volumes in Loughlin et al. (2010):

$$455 \quad \Delta P_{growth}(t) = \rho_{lava} \times g \times \frac{V}{t} \times \frac{1}{\pi \times r_{Dome}^2} \times t = 0.237 \frac{Pa}{s} \times t \quad (4)$$

456 This function is set to start after one hour of the simulation has passed.

457 *Loading of the magmatic plumbing system:* Here, we assume the existence of two stacked
458 magmatic chambers (following Hautmann et al. (2014)) with the lower magma chamber (LMC) at
459 about 13 km depth and the upper magma chamber (UMC) at about 6 km depth (Fig. 4b).

460 The derived pressure build-up in the UMC due to vesicle formation and growth following
461 the July 2003 collapse was about 4 MPa (Fig. 4 in Voight et al. (2006)). Unfortunately, no strain
462 data exist for the May 2006 collapse. Because the removed dome volume during the 2006 collapse
463 was about half that of the 2003 collapse (97 vs 200 Mm³), we assume a pressure increase of
464 $\Delta P_{UMC,col} = 2$ MPa during the 2006 collapse as a first order estimation. This is applied as a
465 boundary load on the UMC, whereby the load is linearly ramped up over the duration of 1 hour.

466 For lack of better data, we apply the simplified equation

$$467 \quad \Delta P = \frac{1}{\beta} \frac{\Delta V}{V} \quad (5)$$

468 for first-order estimation of depressurization values for the UMC and LMC during dome
469 growth, using the volume changes derived by Mattioli et al. (2010) for SHV's third phase of dome
470 extrusion. Gottsmann and Odbert (2014) infer a range of 1 to 10 GPa for the magma bulk modulus;
471 we use $\frac{1}{\beta} = 7$ GPa following Linde et al. (2009). The derived boundary loads are $\Delta P_{UMC,grow} =$
472 -0.4 MPa and $\Delta P_{LMC} = -17$ MPa. Due to the simplifications and the uncertainties in volume
473 changes and the magma bulk modulus, there are large uncertainties associated with these
474 depressurization values, but they suffice for a proof-of-concept model. To take the uncertainties
475 into account, we vary depressurization values in parametric sweeps to test the sensitivity of the
476 model to these values. The third phase of dome extrusion lasted for 627 days; assuming linear

477 deflation, we use the following time-dependent functions that are applied as a boundary load to
478 UMC and LMC, respectively, once the simulation has run for 1 hour (i.e. after dome collapse is
479 over):

$$480 \quad \Delta P_{UMC, grow}(t) = \frac{\Delta P_{UMC, grow}}{627d} \times t \quad (6)$$

$$481 \quad \Delta P_{LMC}(t) = \frac{\Delta P_{LMC}}{627d} \times t. \quad (7)$$

482 The reference collapse model (COL) can be found as an .mph file in online resource 5.

483 Extensive sensitivity studies on input parameters for both lahar and collapse models were
484 performed. Since the wells are less than 2 km from the coastline, and a lateral connection of the
485 aquifer to the open ocean is conceivable, we tested the effect of a constant water table at the lateral
486 aquifer boundary in the lahar models by setting the lateral boundary condition of the aquifer to h
487 = 0 m at all times. We also tested the effect of topography (as shown in Fig. 4b) and a topography-
488 dependent sediment load on the aquifer pressures in the lahar models. To define the sediment load
489 function, we used the difference in pre-eruptive and current valley floor elevation as presented in
490 Froude (2015) (see online resource 6). The explored parameter space and results of these tests can
491 be found in online resources 3, 7 and 8.

492

493 **6 Lahar models: results and discussion**

494 6.1 Results of the reference simulation

495 In the reference simulation of the Belham aquifer, for both the linear (L) and cyclic loading
496 (Lc) models, the simulated hydraulic head rises by about 2 m during the 3 years of increasing

497 sediment load. The cyclic loading is closely mirrored in the aquifer pressure: the hydraulic head
498 rise is strongest during times of largest sediment load (the wet season) and constant during the dry
499 seasons, when no new load is applied (Fig. 5a).

500 For comparison, we also ran the models with a stiffer and less permeable aquifer (models
501 L_s and L_{cs}). Under the same loading conditions, the hydraulic head in the stiffer aquifer rises by
502 only 0.2 m during 3 years of simulation duration. Thus, simulated head changes in the stiffer
503 aquifer are too small by an order of magnitude (Fig. 5a).

504 The main deviation between the reference model and observations in the Belham valley
505 occurs prior to the onset of dome growth (i.e. during the first 212 days). The observed water level
506 increases between January and March 2005, while the model predicts only a small increase in
507 January and a constant well level from February to April. Subsequently, observed well levels fall
508 between June and August 2005, when the model predicts increasing hydraulic heads. After
509 eruption onset, the simulated hydraulic head largely parallels that observed in the Belham aquifer.
510 Minor deviations after eruption onset can still be found during dry seasons, in which observed
511 hydraulic heads fall, while the simulated heads stay constant, and in the period following the dome
512 collapse, where the observed increase in hydraulic head is slightly stronger and less periodic than
513 predicted by our models.

514 6.2 Sensitivity analysis

515 To investigate the influence of input parameters, we performed parametric sweeps
516 regarding aquifer properties, the sediment load and boundary conditions. This section presents the

517 most important results of the sensitivity analysis, additional information can be found in Online
518 Resource 7.

519 Amongst aquifer properties, the Young's Modulus is the most influential parameter. The
520 softer the aquifer, the stronger is the hydraulic head response to the load at the surface (Fig. 5b-c).
521 This matters most in stiffer aquifers: while a decrease of the Young's Modulus from 100 to 0.5
522 MPa has negligible influence, the decrease from 100 to 0.5 GPa increases the final hydraulic head
523 rise from $h \approx 0.1$ to $h \approx 2$ m. Additional sensitivity studies regarding aquifer properties can be found
524 in Online Resource 7.

525 We varied the applied sediment load by changing the assumed sediment thickness and
526 density in the calculation for resulting load. Both parameters have a significant influence on the
527 resulting aquifer pressure; the higher the sediment density and/or the sediment thickness, the
528 stronger the hydraulic head rise (Fig. 5d-e). Within the tested range, a hydraulic head increase of
529 up to $h \approx 8$ m in the soft, and $h \approx 0.8$ m in the stiff aquifer is reached after 3 years.

530 While the hydraulic head in the aquifer is spatially homogeneous in other simulations, it varies
531 laterally in simulations with a constant water table at the lateral aquifer boundary (representing a
532 connection to the ocean): the closer to the “ocean”, the smaller is the total head rise (Fig. 6a-b).
533 This leads to hydraulic head decreases during phases with less or no sediment load due to pressure
534 equilibration in the aquifer, which is facilitated by porous flow towards the boundary. A similar
535 effect occurs in models that incorporate topography and a topography-dependent sediment load on
536 the aquifer pressures: the hydraulic head at the well location decreases during dry seasons as
537 opposed to staying constant in the reference simulation (Fig. 6c-d). This is again due to lateral

538 pressure variations in the aquifer - the hydraulic head is larger closer to the summit due to the
539 higher sediment load - which induces porous flow away from the well location.

540 6.3 Discussion: long-term water level rise by sediment loading

541 Since rain is the main trigger for lahars on Montserrat, the good correlation of cumulative
542 rainfall and water level increases supports the hypothesis of increased (cumulative) lahar deposit
543 loading as a mechanism for the pressure increase in the Belham aquifer. The simulated hydraulic
544 heads in the aquifer induced by sediment loading are of the right order of magnitude and show
545 similar patterns to the observed water level rise. However, there are some clear deviations in the
546 periodicity of the simulated and observed water levels before eruption onset where observed water
547 levels rise during the dry period in early 2005. One possible explanation is that the seasonality of
548 the water level signal prior to eruption onset is dominated by seasonal recharge. The recharge
549 signal in water levels is likely to lag behind the precipitation seasons, thus explaining a rise in
550 water levels with some delay after the wet season. After eruption onset, more loose sediment is
551 available and more lahars enter the valley, thus the seasonal signal now becomes dominated by
552 rainfall seasons. Since our models do not account for recharge, this effect cannot be resolved with
553 this study.

554 In the months directly following the dome collapse in May 2006, the observed increase in
555 water levels is slightly stronger than predicted by our models and lacks the clear seasonality prior
556 to this event. This might be due to the significant shift in sediment deposition after the collapse,
557 which led to heavily loaded lahars even during seasons with usually few or no lahars. This effect
558 was also not included in our models.

559 Several parameters influence how the aquifer responds to the surface loading, but aquifer
560 stiffness and sediment load are the most influential, indicating that the hypothesized deeper and

561 stiffer aquifer is unlikely to contribute significantly to the water level signal due to lahar
562 sedimentation. Boundary conditions come into play in periods representing the dry season, when
563 observed water levels decrease slightly, which according to our models can be explained with a
564 link of the aquifer to the ocean and/or topography-dependent loading. There is definitely some
565 topography-dependent loading as sedimentation in the Belham valley is not spatially
566 homogeneous. This causes pressure imbalances and thereby porous flow in the aquifer from higher
567 to lower pressures, leading to the observed water level decreases in dry seasons.

568 We cannot exclude the possibility of additional factors driving the water level rise in the
569 Belham valley. In particular, as outlined in section 4, an increase in recharge can be a significant
570 driver of rising water levels. Our models do show, however, that the effect of lahar loading is
571 significant, consistent with the observations and causes a non-negligible effect on pore pressures
572 in the aquifer. The detailed sensitivity studies (see also online resource 7) show that this holds true
573 for a wide range of aquifer properties and boundary conditions. Hence, in the analysis of well
574 water levels, the poroelastic effect caused by increasing sediment load must not be disregarded.
575 The periodicity of the signal may be explained with the seasonality of rainfall and thus lahars. But
576 this seasonality can also be seen in hydrological recharge and atmospheric pressure and, as
577 discussed above, there are indications for a seasonal recharge signal in the water level data prior
578 to eruption onset. Since precipitation, atmospheric pressure, lahars and recharge are all inherently
579 coupled, it is not possible to distinguish individual signals with the available data.

580

581 **7 Collapse models: results and discussion**

582 7.1 Results of the reference simulation

583 With parameters at reference values, the temporal evolution of hydraulic head in the stiffer
584 aquifer (COL_s) follows loading due to dome collapse and growth closely: the hydraulic head falls
585 during the collapse to $h \approx -14$ cm and then immediately starts to rise again during renewed dome
586 growth reaching about $h \approx 17$ cm at 200 days (Fig. 7a). The modelled water level fall during dome
587 collapse in the stiff aquifer (COL_s) is similar to the observed value, but occurs simultaneously with
588 dome collapse, while the observed drop has a time delay of 6-11 days after the collapse (there are
589 no data points available between the 25th and 31st of May). The response of the stiff aquifer to
590 dome growth-related strain is smaller than observed water level increases by a factor of 4.

591 The hydraulic head in the simulated less competent Belham aquifer (COL) falls by about
592 0.05 cm during dome collapse. It then continues to fall until about 20 days after the collapse to
593 reach a minimum of $h \approx -0.15$ cm, which is two orders of magnitude less than the observed values.

594 Loading due to dome growth then leads to a head rise, reaching $h \approx 0.1$ cm at the end of the
595 simulation, i.e. at 200 days (Fig. 7a).

596 7.2 Sensitivity analysis

597 To investigate the influence of input parameters, we performed parametric sweeps
598 regarding aquifer properties and magma chamber loads. Here, we show only the most important
599 results, additional sensitivity studies can be found in Online Resource 8.

600 Results of parametric sweeps on the aquifer's Young's Modulus (Fig. 7b-c) indicate that
601 the difference in stiffness between the two types of aquifers is the main cause for the differing
602 hydraulic head changes in response to dome collapse. Generally, the larger the Young's Modulus,
603 the larger are the decreases in hydraulic head following dome collapse and the increases in
604 hydraulic head during dome growth. Increasing the Young's Modulus of the soft aquifer to 100
605 MPa increases the head fall to $h \approx -1$ cm; the largest head fall is reached for $E=50$ GPa with $h \approx -14$
606 cm. Additionally, the time between dome growth onset and onset of hydraulic head rise is shorter
607 the larger the Young's Modulus.

608 Aquifer depth also significantly affects the resulting hydraulic head fall in the COL model
609 suite. At a depth of 1.5 km, the simulated collapse-related fall in hydraulic head in the Belham
610 aquifer is $h \approx -11$ cm (Fig. 7d). In the stiffer aquifer, hydraulic head fall is smallest for an aquifer
611 depth of 1 km, and is larger for shallower or deeper aquifers (Fig. 7e). However, the influence of
612 aquifer depth on changes in hydraulic head in the stiff aquifer is small in comparison with other
613 parameters.

614 7.3 Discussion: short-term water level fall by dome collapse-related strain

615 As expected (see section 4), simulated strains related to dome growth and collapse cause a
616 very small response in the reference Belham aquifer. Simulated hydraulic heads fall to their
617 minimum values with a time delay of several days after the dome collapse, but the fall is two orders
618 of magnitude smaller than observed. Many parameters influence this result (see also online
619 resource 8). However, the only model which produces the observed magnitude of hydraulic head
620 decrease in a soft aquifer invokes a 1.5 km deep aquifer (Fig. 7d). The observation wells tap a soft,
621 shallow aquifer, which therefore, according to our models, would not respond to dome-collapse
622 related strains.

623 Since it responds to sediment load, the Belham aquifer would also respond to erosion, i.e.
624 the removal of sediment burden. The long-term trend in the valley is sediment aggradation but
625 during individual events or episodes, local erosion commonly occurred in the form of channel
626 incision (Alexander et al., 2010, Froude, 2015). We therefore tested whether this could be an
627 alternative explanation for the water level fall after the May 2006 dome collapse, which was
628 followed by a period of intense lahars. But the simulated channel incision only causes a sufficiently
629 large hydraulic head drop if there is large scale (at least 50 cm deep), almost valley-wide (50-100
630 m wide) erosion (see online resource 9), which did not occur.

631 However, dome collapse-related strain produces the right order of magnitude of hydraulic
632 head fall in a much stiffer and less permeable aquifer. As expected for a poroelastic medium, the
633 simulated response is instantaneous, while the observed response is delayed by at least 6 days. The
634 hydraulic heads in this aquifer increase in response to dome growth, to about 25% of observed
635 values.

636 Model results for a stiff aquifer hence support the hypotheses that the shallow Belham
637 aquifer is connected to a second, stiffer aquifer and that the significant pressure drop due to dome-

638 collapse in that stiffer aquifer leads to water leakage from the Belham aquifer. This would then
639 cause well levels in the Belham aquifer to drop with some delay. Hemmings et al. (2015a, 2015b)
640 have already proposed the existence of a deeper aquifer in the Belham valley. As discussed in
641 section 4, an increase in rock density and competence at greater depths is reasonable and it is likely
642 that the deeper-seated aquifer is connected with the Belham aquifer by vertical fractures. The
643 connection between the aquifers may be intermittent or permanent. If the Belham aquifer is fully
644 connected to a stiffer aquifer at all times, the pressure increase in the stiffer aquifer due to dome
645 growth could be contributing to the observed hydraulic head rise. The well level rise can be
646 explained solely with lahar loading, but parameter uncertainties in our models are such that we
647 cannot discard processes with certainty.

648 For a first-order investigation, we ran an additional simple 2D simulation, incorporating
649 two stacked aquifers that are separated by an aquitard, but connected by vertical fractures (see
650 online resource 10). Initialising the deeper aquifer at a hydraulic head of -0.1m indeed leads to
651 downward fluid flow through the fractures and a hydraulic head fall in the upper aquifer reaching
652 -0.09 m after 11 days with the chosen parameters. This further confirms the feasibility of our
653 hypothesis, although the uncertainties regarding geometry and material properties are significant.
654 More sophisticated modelling and detailed parameter studies of such a two-aquifer model, in
655 particular one that also incorporates the lahar loading, is beyond the scope of this study and left
656 for future work.

657 No strain data exist for the time period considered here. Data series from the GPS network
658 show no clear indication for ground deformation at proximal sites and no evidence for any
659 deformation in the Belham Valley associated with the dome collapse in May 2006 (Pascal et al.,
660 2017). Under the assumption that the well level changes are indeed caused by dome collapse-

661 related strain, we can utilise these to draw implications about crustal strain. We applied two strain
662 sources during dome collapse in our simulations: pressure build-up of the UMC and unloading of
663 the summit. Both sources can cause a hydraulic head fall of similar magnitude and most likely
664 acted together (see online resource 11). Therefore, the water level record indicates a pressure build-
665 up in the upper magma chamber during a dome collapse, supporting the conclusions drawn from
666 strainmeter data in 2003 by Voight et al. (2006).

667

668 **8 Conclusions**

669 Volcanic strain has been the suggested driver behind water level changes at many
670 volcanoes, but this process has so far not been sufficiently quantitatively tested. Strehlow et al.
671 (2015) developed numerical models to investigate a hypothetical scenario of poroelastic aquifer
672 responses to magma chamber inflation. This study has now extended these efforts by adapting the
673 models to a real field scenario with specific water level observations, illustrating that poroelastic
674 effects are not purely theoretical but can play an important role in real monitoring data.

675 Even though the results do not perfectly match the observations, the presented simulations
676 can help to decipher origins of observed water level changes in the Belham valley on Montserrat,
677 that comprise both a long-term well level rise over the period 2004-2006 and a short-term water
678 level fall in May 2006. Simulated pore pressures in the Belham aquifer increase significantly due
679 to the long-term sediment deposition by repeated lahars in the valley, which in turn leads to a well
680 water level rise. While not notably affecting the Belham aquifer itself, simulated crustal strains
681 associated with the dome collapse in May 2006 lead to a significant water level drop in a more
682 competent aquifer. Thus, water could leak from the Belham aquifer e.g. through connecting

683 fractures, into a deeper-seated, stiff aquifer, thereby causing a noticeable well level drop days after
684 the collapse.

685 Therefore, the suggested conceptual model shown in Fig.3 represents one possible
686 explanation for the water level observations between 2004 and 2006: We suggest that both
687 sediment accumulation and dome collapse significantly affected the hydraulic head in the Belham
688 aquifer on different timescales. While additional hydrological drivers cannot be excluded with
689 absolute certainty, the suggested processes can serve as a possible explanation for the observed
690 water level variations and cannot be neglected in the analysis of well water levels.

691 Our study shows that groundwater dynamics in relation to volcanic activity as predicted
692 from theoretical modelling (Strehlow et al. 2015) can indeed be witnessed in field observation data
693 and that volcanic strain from various sources strongly affects water levels in regional aquifers on
694 different time scales. Therefore, this study represents an intermediate step between the theoretical
695 analysis in Strehlow et al. (2015) and the incorporation of water level observations in volcano
696 monitoring systems. The inferred underlying processes are not unique to the island of Montserrat
697 and as such make our results transferable to other volcanic and hydrological settings, where the
698 observation and interpretation of water levels can also provide important insights regarding the
699 state of volcanoes.

700

701 8.1 Future work

702 While our study provides a possible first-order explanation for observed water level
703 variations in the Belham valley, there are still deviations between the model results and
704 observations and our models include several simplifications and unknowns. To fully exploit the
705 potential of water level observations within volcano monitoring systems, both further development

706 of the numerical models and more sophisticated hydrological data acquisition and analysis are
707 necessary. We recommend:

- 708 • A quantification of seasonal and long-term variations in hydrological recharge, and
709 incorporation into the models.
- 710 • The development of full three-dimensional models to truly incorporate topographic and
711 stratigraphic features of Montserrat both on-land and offshore.
- 712 • The development of a stacked two-aquifer model that combines the effects of strain due to
713 lahar loading and dome collapse in one simulation.
- 714 • The set-up of inversion models to acquire a best-fit model to the observations.
- 715 • (For any volcano:) The routine acquisition of high-quality, high-resolution water level data,
716 in order to resolve barometric and tidal effects on the aquifer. Since tidal strains are known,
717 the response to tides can then be used to calibrate water wells as strainmeters and help to
718 ground-truth numerical models (e.g. Roeloffs 1996; Wang 2000).

719

720 **References**

721 Availability of data and models: All data used in this study, as well as reference models (.mph
722 files to be opened in COMSOL Multiphysics) for both the lahar and the collapse models can be
723 found in the supplementary materials online (online resources 4, 5 and 12).

724

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935 **Tables**

936 **Table 1:** Correlation coefficients of different data sets with water levels in the Belham valley wells. Size

937 of the water level data set is n=92

Data set	Correlation coefficient	p-Value	Size of dataset
Precipitation (daily)	-0.02	0.57	791
Precipitation (cumulative)	0.96	$< 10^{-9}$	791
Mean sea level pressure (daily)	0.13	4×10^{-4}	784
Mean sea level pressure(daily) – pre-collapse	0.28	$< 10^{-4}$	563
Mean sea level pressure (daily) – post-collapse	-0.33	$< 10^{-4}$	221
Mean sea level pressure (monthly)	-0.26	0.21	26
Mean sea level pressure (monthly) – pre-collapse	0.9	$< 10^{-4}$	19
Mean sea level pressure (monthly) – post-collapse	0.25	0.54	7

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939 **Table 2:** Reference values of input parameters. Abbreviations: B-aq = Belham aquifer, S-aq = Stiff
 940 aquifer, BW-coeff. = Biot-Willis-coefficient, Young's = Young's Modulus, Cap = the impermeable layer
 941 above the aquifer

Parameter	Symbol & reference value	Parameter	Symbol & reference value
B-aq BW-coeff.	$\alpha = 0.7$	S-aq BW-coeff.	$\alpha = 0.3$
B-aq permeability	$\kappa = 1 \times 10^{-10} \text{ m}^2$	S-aq permeability	$\kappa = 1 \times 10^{-14} \text{ m}^2$
B-aq porosity	$\phi = 0.35$	S-aq porosity	$\phi = 0.1$
B-aq Young's	$E_{aq} = 10 \text{ MPa}$	S-aq Young's	$E_{aq} = 50 \text{ GPa}$
B-aq density	$\rho_{aq} = 2000 \text{ kg/m}^3$	S-aq density	$\rho_{aq} = 2800 \text{ kg/m}^3$
Aquifer thickness, L	$d_{aq} = 50 \text{ m}$	Lateral distance	$L = 3 \text{ km}$
Aquifer thickness, COL	$d_{aq} = 200 \text{ m}$	Aquifer depth	$z_{aq} = 50 \text{ m}$
Crust density	depth-dependent _a	Water compressibility	$\chi = 4 \times 10^{-10} \text{ Pa}^{-1}$
Crust Young's	depth-dependent _b	Water density	$\rho_f = 1000 \text{ kg/m}^3$
Cap density	$\rho_{cap} = 2000 \text{ kg/m}^3$	Water viscosity	$\mu = 10^{-3} \text{ Pas}$
Cap Young's	$E_{cap} = 10 \text{ MPa}$	Sediment aggradation	$a = 0.4 \text{ m/a}$
Poisson's ratio	$\nu = 0.3$	Magma bulk modulus	$\frac{1}{\beta} = 7 \text{ GPa}$
UMC centre depth	$z_{UMC} = 6 \text{ km}$	LMC centre depth	$z_{LMC} = 12.5 \text{ km}$
UMC depressurization	$\Delta P_{UMC, grow} = -0.4 \text{ MPa}$	LMC depressurization	$\Delta P_{LMC} = -17 \text{ MPa}$
UMC radius	$r_{UMC} = 1.25 \text{ km}$	LMC vertical semi-axis	$b_{LMC} = 2 \text{ km}$
UMC pressure increase _c	$\Delta P_{UMC, col} = 2 \text{ MPa}$	LMC horizontal semi-axis	$c_{LMC} = 1.7 \text{ km}$
Sediment density	$\rho_s = 1800 \text{ kg/m}^3$		

942 ^aBecause the crust gets denser with depth, we use $\rho_c = 2800 \text{ kg/m}^3$ for depths $<2 \text{ km}$ and a ρ_c -vs-depth
943 function for greater depths as defined in Gottsmann and Odbert [2014] and Young and Gottsmann [2015].

944 ^bBecause the crust gets stiffer with depth, we use $E_c = 5 \text{ GPa}$ for depths $<2 \text{ km}$ and a E -vs-depth function
945 for greater depths as defined in Gottsmann and Odbert [2014] and Young and Gottsmann [2015].

946 ^cduring dome collapse

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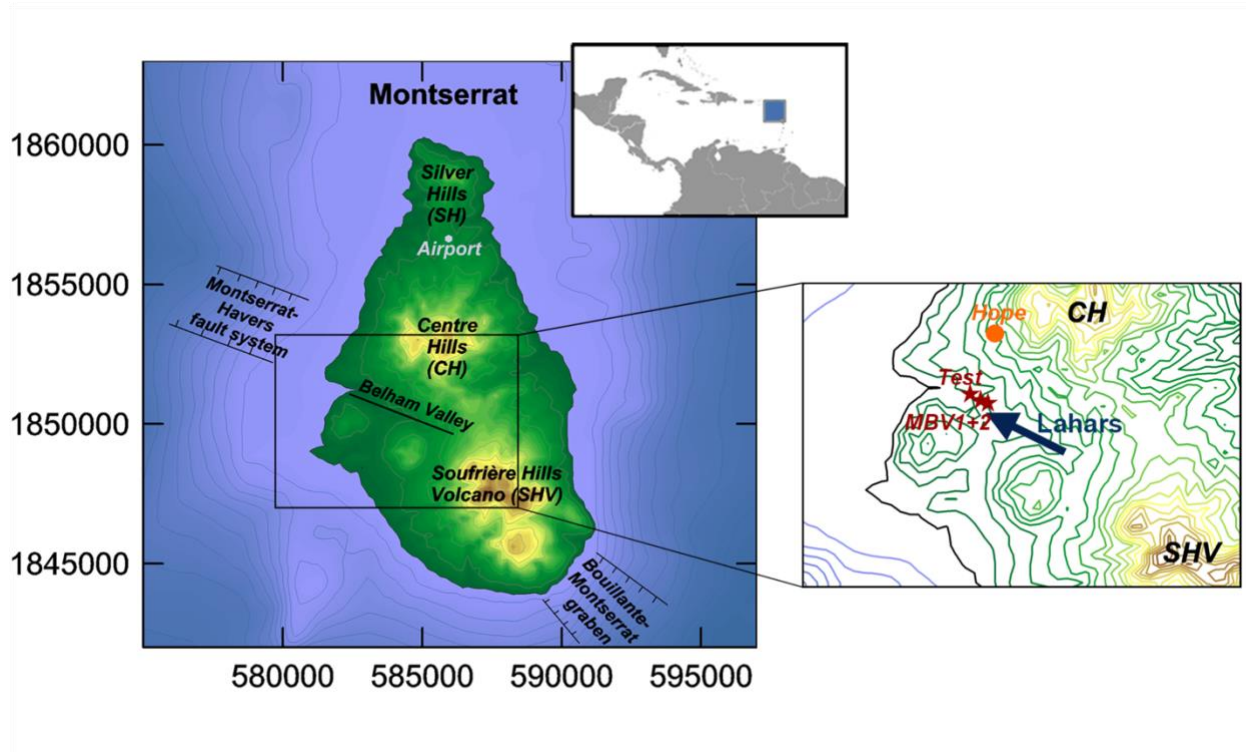
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959 **Figures**



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961 **Fig. 1:** Overview map of Montserrat and a zoom in on the Belham valley, with indicated well locations
962 (MBV1, MBV2 and Test), the three volcanic centres, lahar flow directions, Montserrat's airport and the
963 Hope rain gauge.

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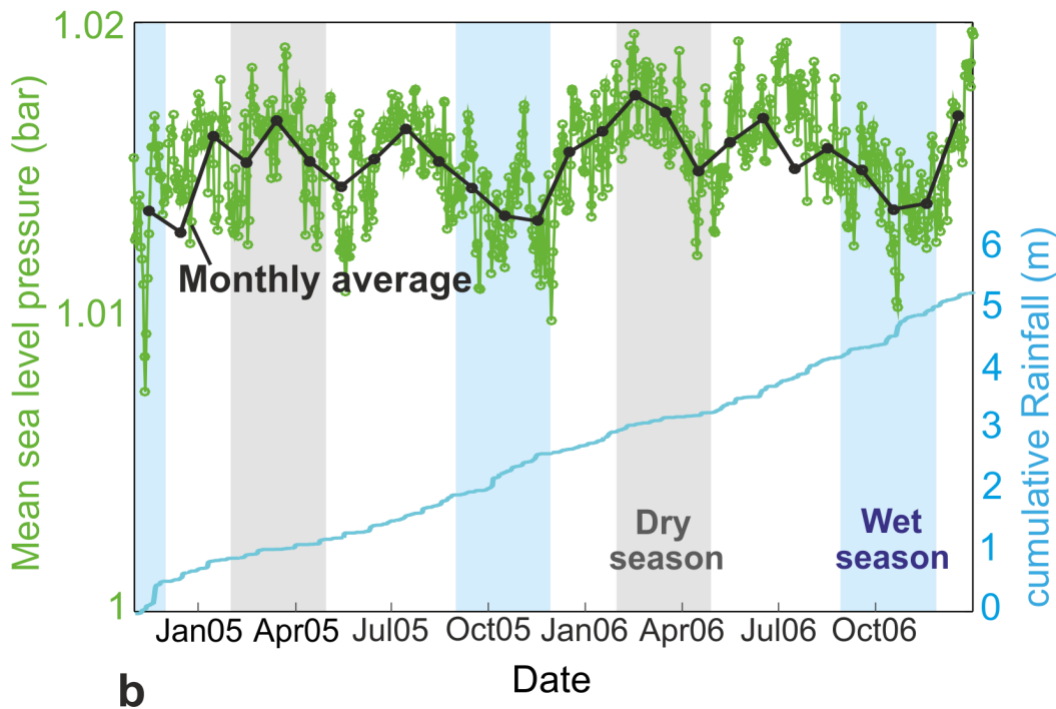
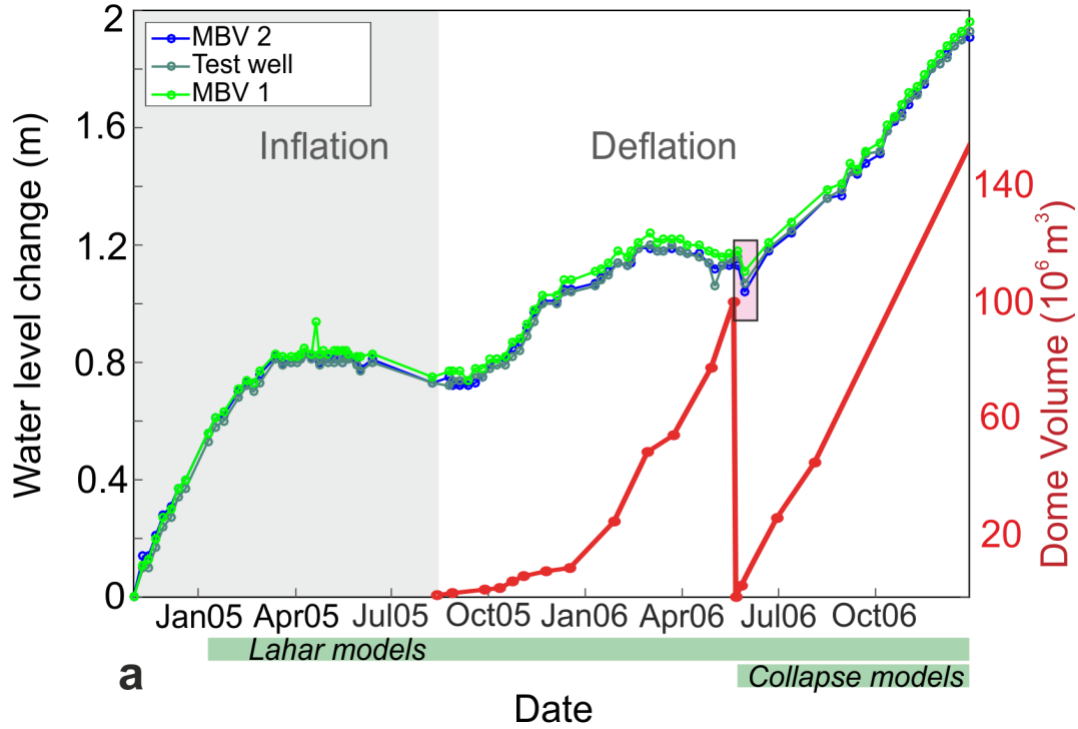
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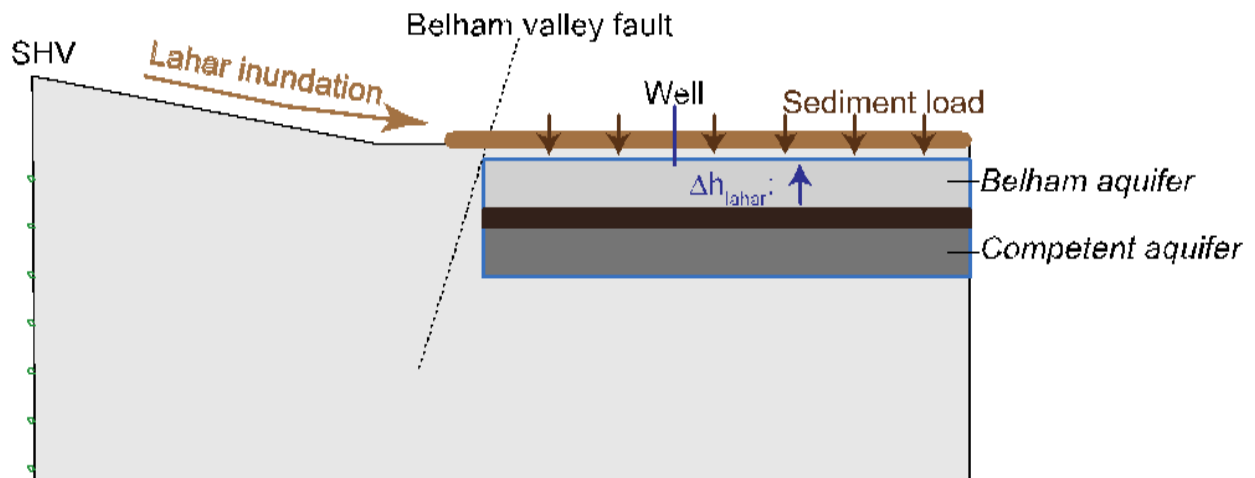
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982 **Fig. 2:** (a) Water level changes in the three observation wells in the Belham valley (owned by Montserrat
983 Utilities) and dome volume after Ryan et al. (2010). The final depth to water level was 3.87 m for MBV

984 1, 1.29 m for MBV 2 and 4.7 m for the test well. Water level data can be found in online resource 12.
985 Inflation and deflation-phases of SHV's edifice are indicated, as well as the modelled water level fall in
986 May 2006 (pink box) and the simulation periods for both lahar and collapse models (green boxes). (b)
987 Cumulative rainfall data (owned by Montserrat Utilities, from the Hope rain gauge (see Fig. 1)) and
988 mean sea level pressure at Montserrat's airport (data provided by Karen Pascal, owned by the
989 Government of Montserrat). Rainfall and sea level pressure data can be found in online resource 12. Wet
990 and dry periods are indicated.

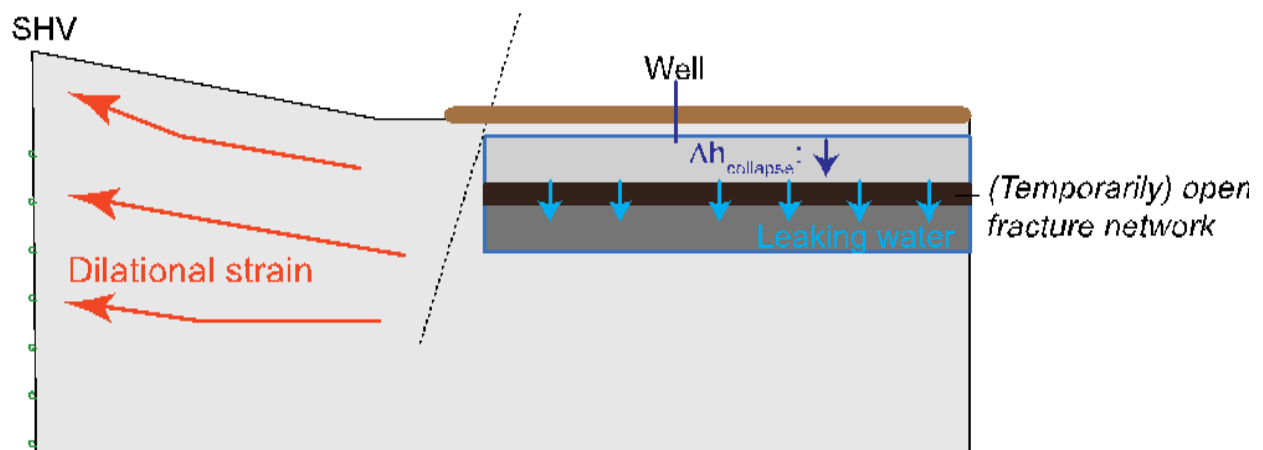
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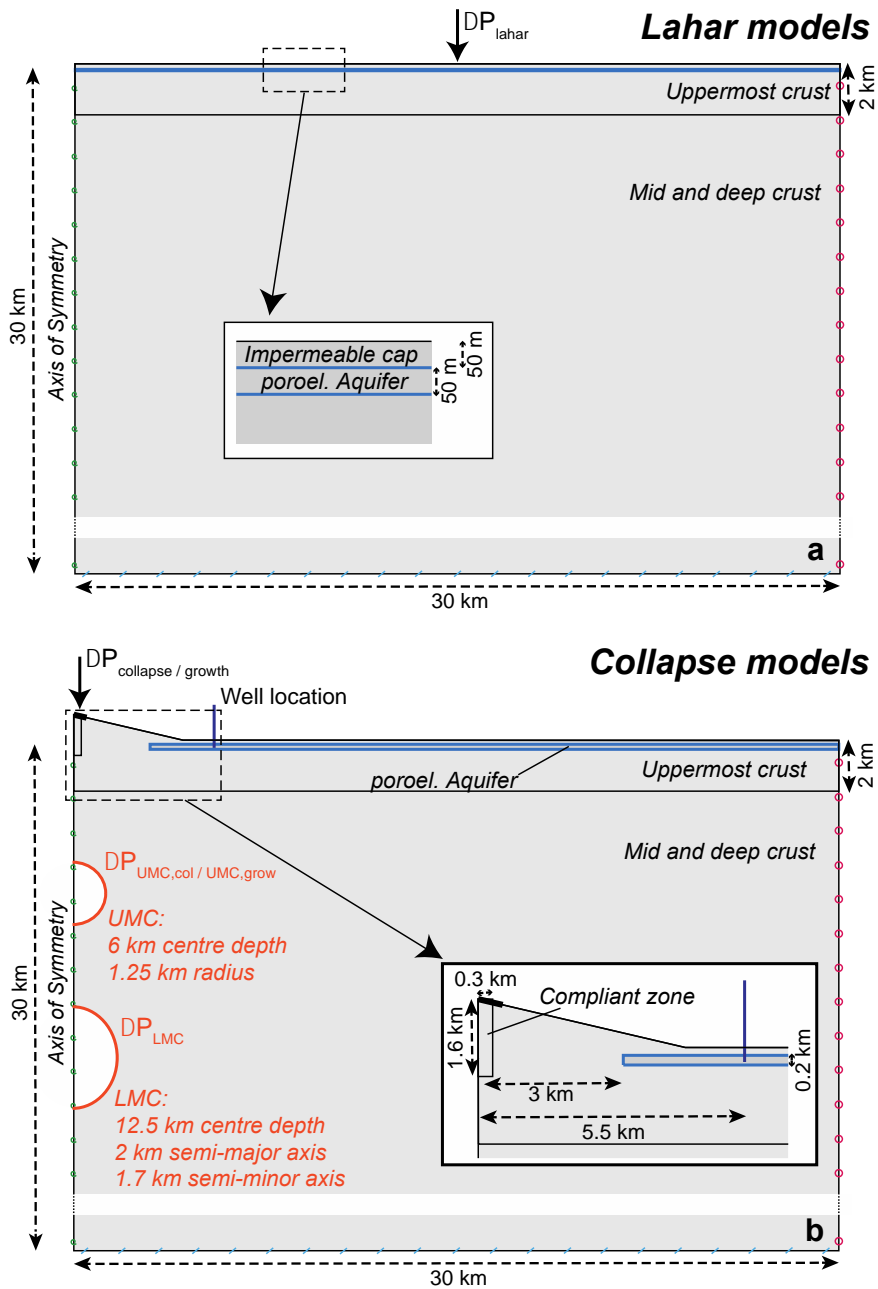
Background situation: Repeated lahar sedimentation



Transient situation: Dome collapse related strain



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994 **Fig. 3:** Conceptual model / hypotheses for hydraulic head changes in the Belham aquifer. The general
995 background situation involves pressure increase in the Belham aquifer due to lahar sedimentation in the
996 valley and a consequent increase in surface load. A transient occurred during the dome collapse, when
997 unloading of the summit and pressure build-up in the UMC led to dilatational strains that caused hydraulic
998 heads to fall significantly in a deeper-seated, competent aquifer. This led to water leaking from the Belham
999 aquifer through a (possibly temporary) fracture network into the deeper system.



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Fig. 4: The lahar (a) and collapse models (b). Topographic height of the summit is 1000 m.a.s.l., radius of the loaded area on the summit is 500 m. Lateral boundary has a roller condition (free lateral, no vertical displacement), bottom boundary is fixed. In (a) the whole surface is loaded according to lahar sedimentation as given in equation 2; in (b), the surface is free. The boundary conditions bordering the aquifer domain are (i) no flow and (ii) continuous stress and displacement. In the dome collapse model, a more compliant zone in the centre of the summit is included with $E=0.5$ GPa, following Young and Gottsmann (2015). Within

1007 the different layers, material properties are considered isotropic and homogeneous. Groundwater flow is
1008 limited to the aquifer domain. Well location in models including topography is indicated in (b). In models
1009 without topography, hydraulic head changes are evaluated in the centre of the domain (and are the same
1010 everywhere in the model) unless otherwise noted. The magma chambers UMC and LMC are represented
1011 by loaded cavities in the domain in (b) (indicated in red).

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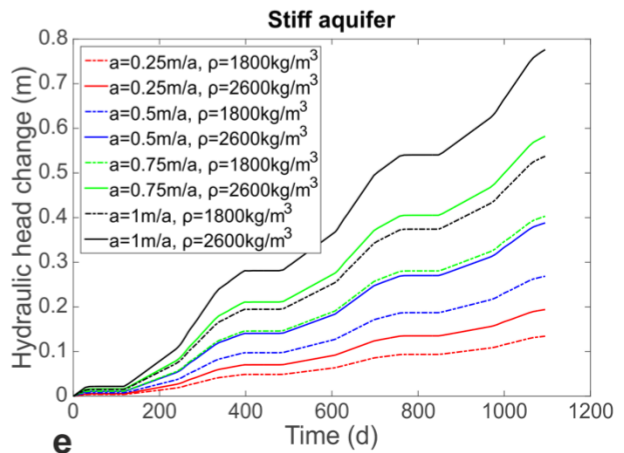
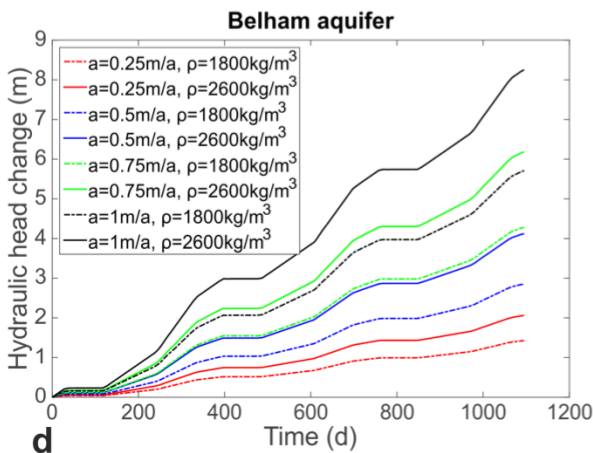
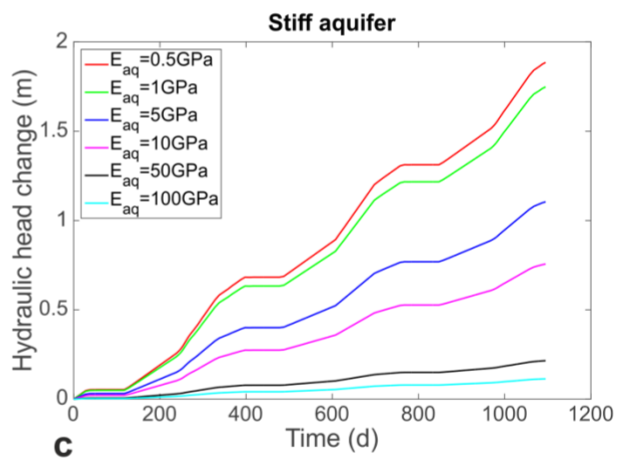
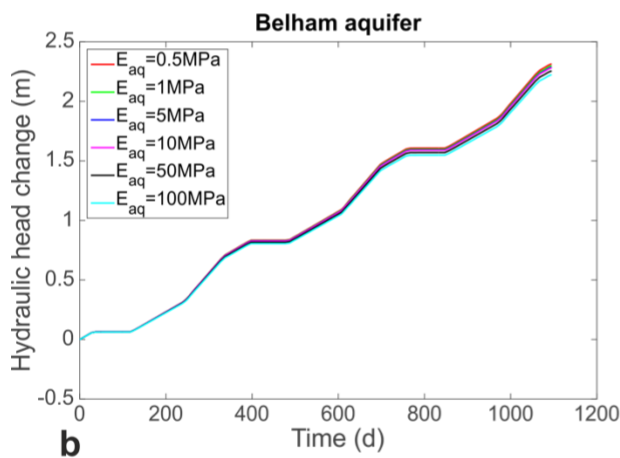
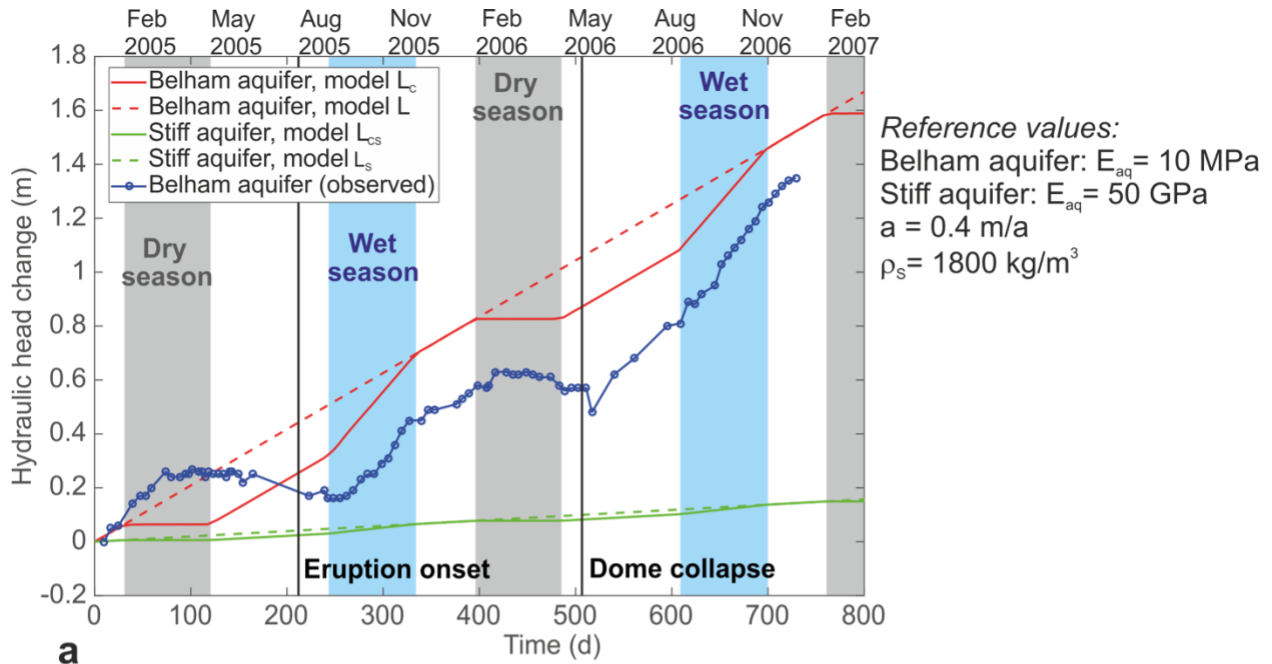


Fig. 5: (a) Hydraulic head changes with time calculated in different models for a long-term water level rise by sediment loading (both linear and cyclic loading applied to the reference Belham aquifer (models L and L_C) and a stiffer aquifer (models L_{Cs} and L_S)), in comparison to observed changes in the water level of MBV 2 from January 2005 onwards. Wet and dry seasons, as well as the timing of eruption onset and dome collapse are indicated. (b)-(e) show results of selected sensitivity studies on Lahar models with cyclic loading: Hydraulic head change with time for a varying Young's Modulus of the aquifer E_{aq} (b, c) and varying sediment density ρ and aggradation a (d, e).

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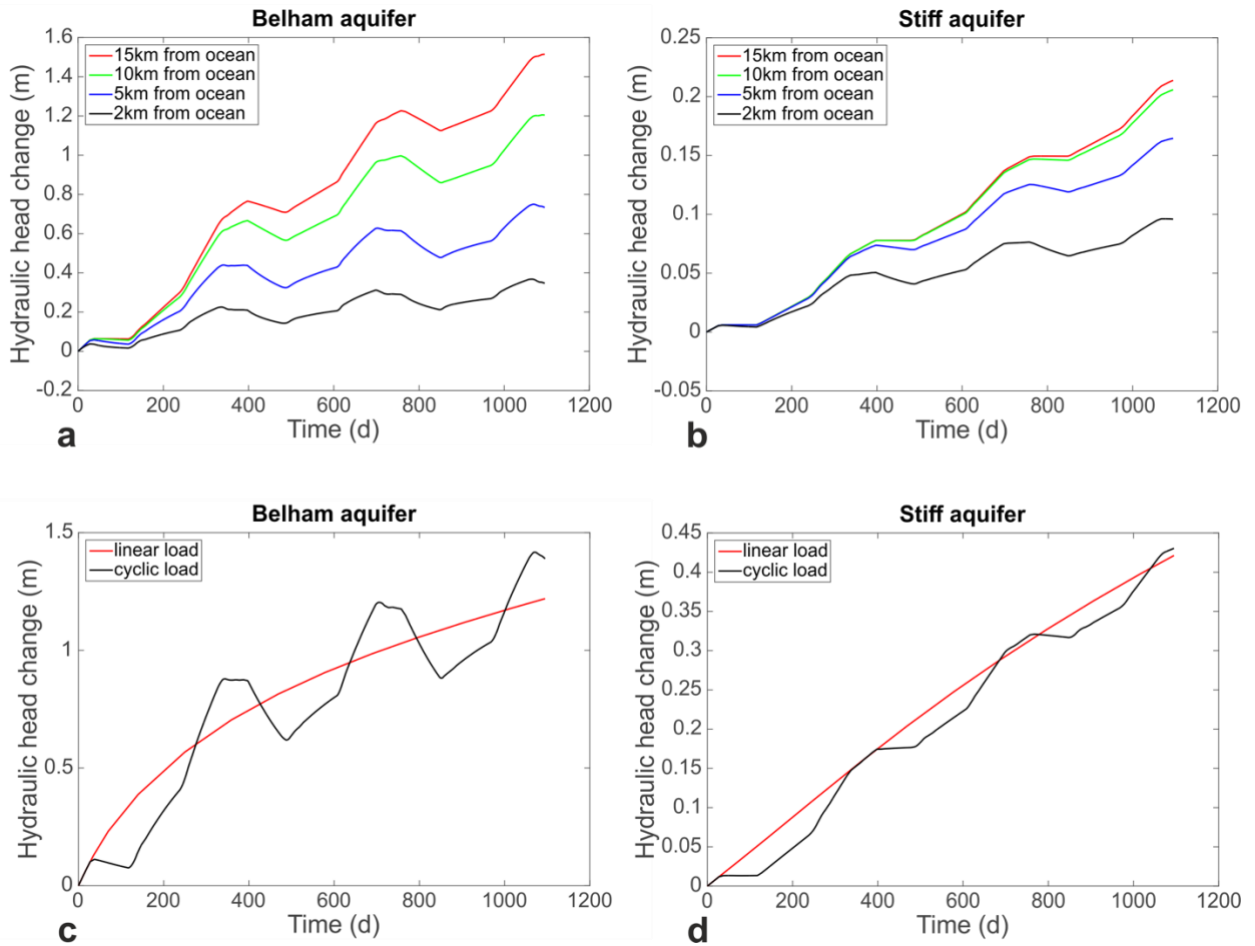
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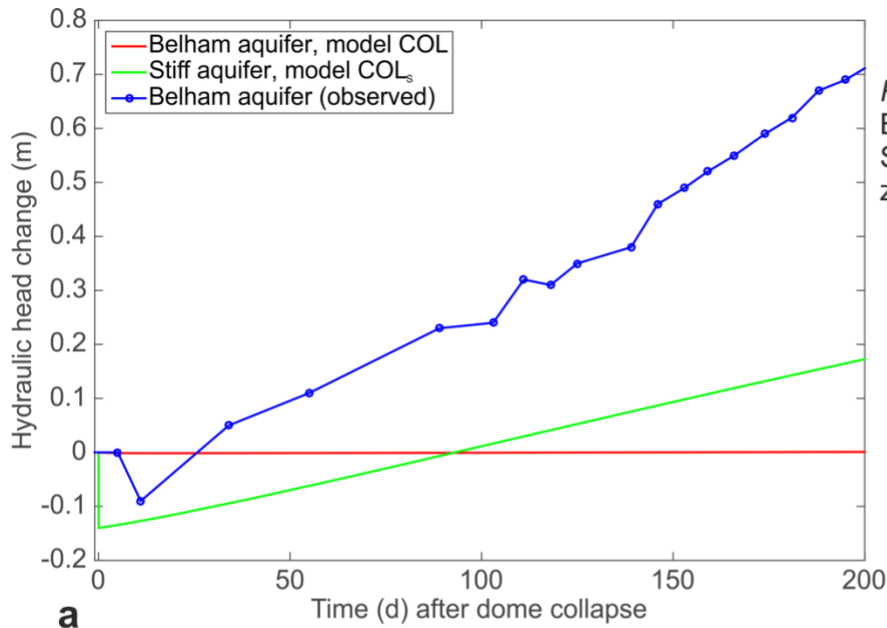
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Fig. 6: Results of selected sensitivity studies on Lahar models: Hydraulic head change with time for different distances from the ocean, with a constant water table as lateral boundary condition and cyclic loading (a, b); hydraulic head change with time for topography-dependent surface loading, showing both linear and cyclic loading (c, d).

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Reference values:
 Belham aquifer: $E_{aq} = 10 \text{ MPa}$
 Stiff aquifer: $E_{aq} = 50 \text{ GPa}$
 $z_{aq} = 50 \text{ m}$

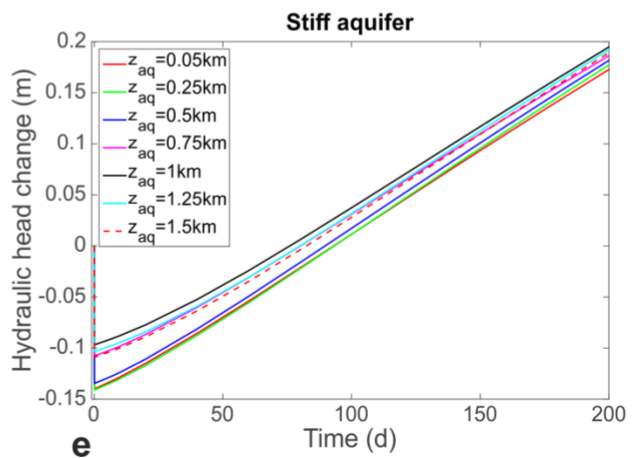
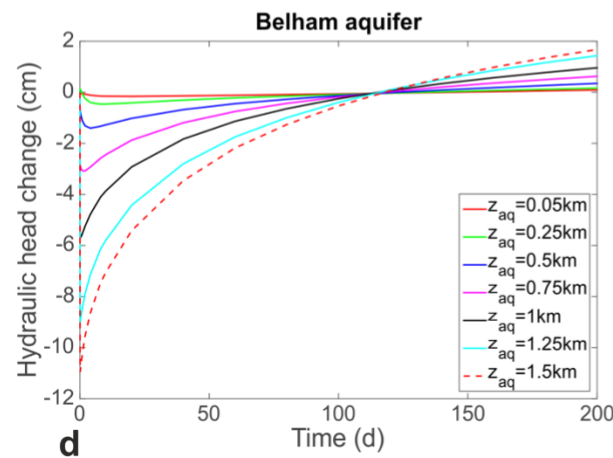
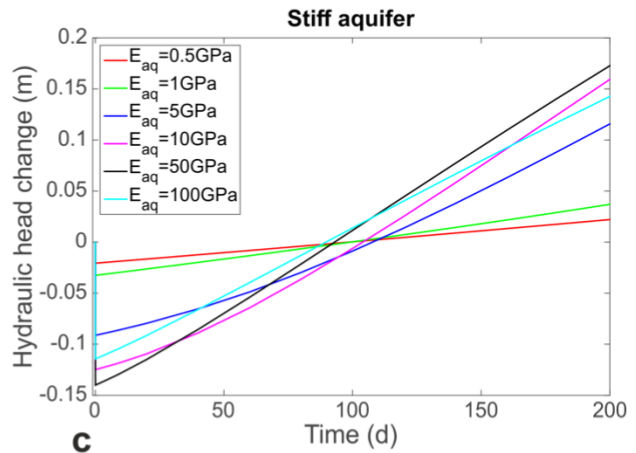
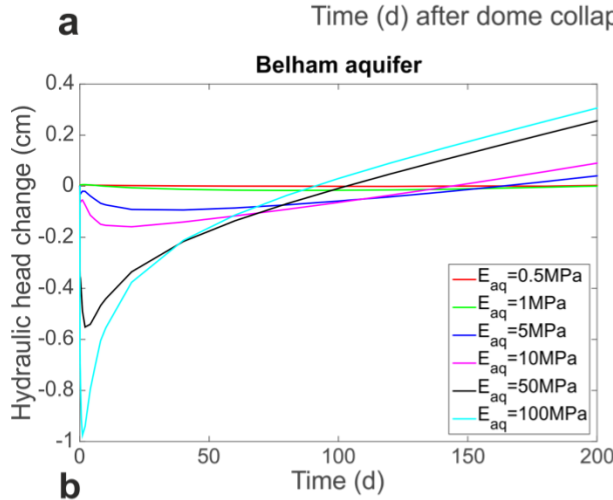


Fig. 7: (a) Hydraulic head changes with time calculated in models for the water level response to dome growth and collapse (for the reference Belham aquifer (model COL) and a stiffer aquifer (model COLs)), in comparison to observed changes in the water level of MBV 2 from the moment of dome collapse onwards. (b)-(e) show results of selected sensitivity studies: hydraulic head change with time for a varying Young's Modulus of the aquifer E_{aq} (b, c) and aquifer depth z_{aq} .

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